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Evidence for Cenozoic tectonic deformation in SE Ireland and near offshore

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[1] An integrated study of topography, bathymetry, high-resolution aeromagnetic data, and structural observations demonstrates significant Cenozoic fault activity in SE Ireland. Tectonically generated knickpoints and reddened fault breccias along topographic escarpments that are underlain by greywacke bedrock and trend oblique to the regional Caledonian strike provide evidence for fault displacement. Near-offshore faults with similar geometry produce present-day bathymetric scarps and localized tectonic topography in the inverted Kish Bank Basin. The integration of offshore high resolution aeromagnetic data and structural interpretation of the Kish Bank Basin provides evidence for dextral transtension on NNW trending faults and sinistral transpression on ENE trending faults bounding a lower Paleozoic to Carboniferous basement block. These faults correlate onshore with previously recognized Caledonian faults producing topographic offsets and surface uplift. To the north, offshore structures can be traced onshore and cut exposures of the Caledonian Leinster Granite. Structural analysis of these outcrops indicates post-Variscan deformation. A major fault on the NW margin of the batholith cuts a major erosion surface developed on Carboniferous carbonate rocks, and nonmarine Miocene deposits are preserved above this surface. Fault kinematics provide evidence of two paleostress systems: (1) NW-SE σ_1 , subvertical σ_2 and NE-SW σ_3 followed by (2) a clockwise swing of σ_1 to NNW-SSE. Timing of deformation in both stress systems is probably post-Oligocene age. The mechanism driving this deformation is likely ridge-push. Much is already known about Cenozoic tectonics and exhumation from offshore basins. This study shows that onshore Ireland has also been

affected by significant tectonic activity and exhumation during the Cenozoic. **INDEX TERMS:** 8010 Structural Geology: Fractures and faults; 8030 Structural Geology: Microstructures; 8110 Tectonophysics: Continental tectonics—general (0905); 8168 Tectonophysics: Stresses—general; 9335 Information Related to Geographic Region: Europe; **KEYWORDS:** Ireland, Cenozoic, reactivation, uplift, knickpoints. **Citation:** Cunningham, M. J. M., A. W. E. Phillips, and A. L. Densmore (2004), Evidence for Cenozoic tectonic deformation in SE Ireland and near offshore, *Tectonics*, 23, TC6002, doi:10.1029/2003TC001597.

1. Introduction

[2] It is becoming increasingly apparent from recent studies that much of the NW European margin has experienced substantial vertical movements in Cenozoic time [e.g., *Japsen and Chalmers*, 2000; *Doré et al.*, 2002]. A major global sea level regression began at the end of the Cretaceous [*Haq et al.*, 1987], followed by an episode of regional, thermally driven rock uplift linked to imposition of the Iceland plume, and opening of the contemporary North Atlantic Ocean at ~53 Ma [e.g., *Nadin et al.*, 1997; *Doré et al.*, 1999]. The exhumation associated with this thermal event affected, to varying degrees, most areas of NW Europe, [e.g., *Lovell and White*, 1997; *Jones et al.*, 2002].

[3] Significant episodes of rock uplift and deformation continued from late Eocene to Miocene time [e.g., *Japsen and Chalmers*, 2000; *Japsen et al.*, 2002; *Rohrman et al.*, 2002]. This includes inferred post-early Eocene fault reactivation in SE Ireland [*Cunningham et al.*, 2003], syn- and post-Oligocene inversion around the Irish-British Atlantic margin [*Roberts*, 1989], and elongate inversion domes offshore central Norway [*Doré et al.*, 1999]. Significant doming in Scandinavia also occurred in the Neogene, followed and enhanced by rapid Plio-Pleistocene glacial erosion and isostatic rebound during interglacial periods [*Riis and Fjeldskaar*, 1992; *Doré et al.*, 1999].

[4] Many aspects of this Cenozoic activity however are poorly understood due to a discontinuous and poor Cenozoic stratigraphic record. This is particularly apparent on the Irish landmass where, apart from some scattered outcrops of Cenozoic sediment across the island [*Drew and Jones*, 2000], most of the direct onshore evidence for Cenozoic tectonic activity comes from NE Ireland (Figure 1). The

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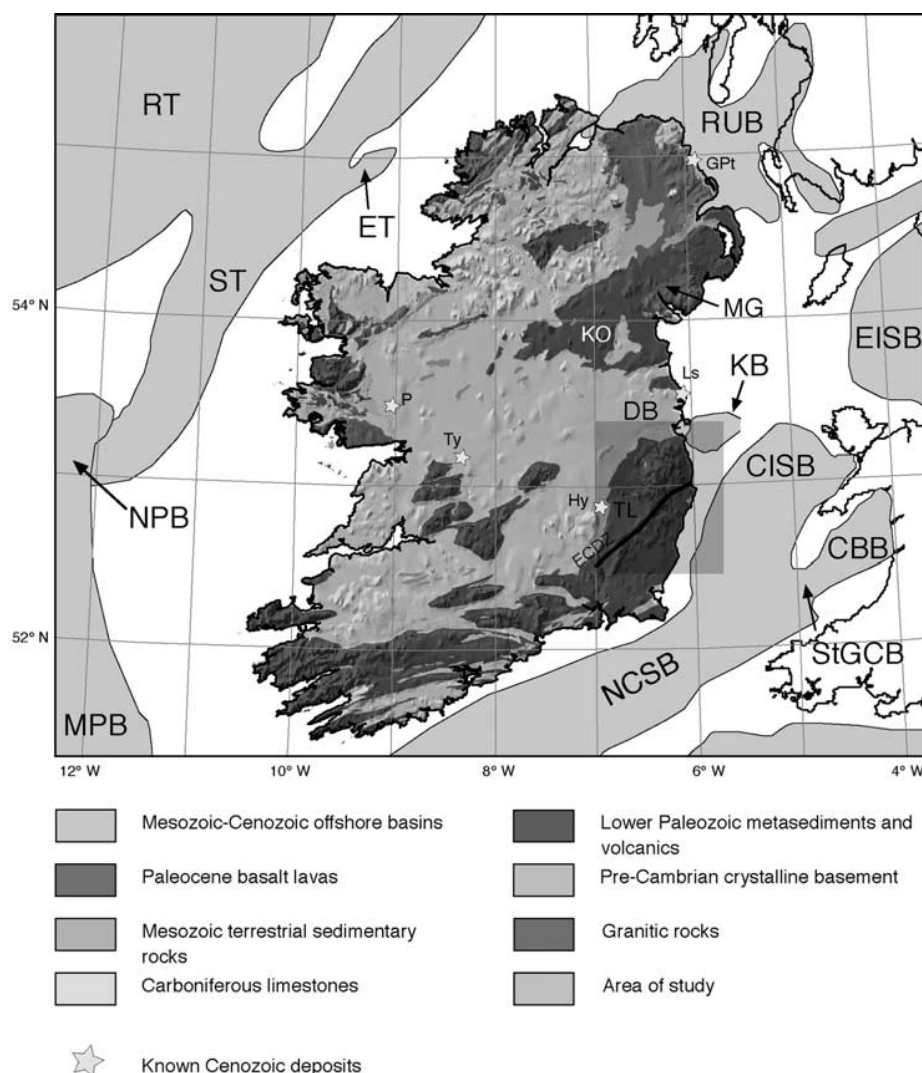


Figure 1. (a) Onshore geology and topography of Ireland and major Mesozoic-Cenozoic offshore basins. Study area is shown by the gray box in SE Ireland. Abbreviations area as follows: CBB, Cardigan Bay Basin; CISB, Central Irish Sea Basin; DB, Dublin Basin; ECDZ, East Carlow Deformation Zone; EISB, East Irish Sea Basin; ET, Erris Trough; Gpt, Garron Point; Hy, Hollymount; KO, Kingscourt Outlier; Ls, Loughshinney; MG, Mourne Granite; MPB, Main Porcupine Basin; NCSB, North Celtic Sea Basin; NPB, North Porcupine Basin; P, Pollnahallia; RUB, Rathlin/Ulster Basin; RT, Rockall Trough; StGCB, St. George's Channel Basin; ST, Slyne Trough; TL, Tullow Lowland; Ty, Tynagh. See color version of this figure at back of this issue.

Paleocene–early Eocene thermal event is associated with emplacement and displacements of dike swarms and sills mainly in the north and west of Ireland and extrusive volcanism in the NE [Gibson *et al.*, 1987; Mussett *et al.*, 1988; Meighan *et al.*, 1999]. During the Oligocene, dextral transtension of NW trending faults in NE Ireland and the Irish Sea (e.g., the Newry–Codling Faults, Figure 1) resulted in the creation of the Lough Neagh-Ballymoney pull-apart basins [George, 1967; Parnell *et al.*, 1988; Naylor, 1992; Geoffroy *et al.*, 1996; Meighan *et al.*, 1999]. The general picture that emerges from the onshore

stratigraphic record and recent apatite fission track studies [Green *et al.*, 2000; Allen *et al.*, 2002; Cunningham *et al.*, 2003] is of a broad patchwork of repeated episodes of exhumation, mostly restricted to the high areas around the margins of the island, with a low-relief erosion surface occupying much of central Ireland (Figure 1).

[5] On the basis of vitrinite reflectance profiles [e.g., Corcoran and Clayton, 2001], apatite fission track analysis [Lewis *et al.*, 1992; Green *et al.*, 1997] and sonic velocity logs [Ware and Turner, 2002], the Irish Sea region experienced 1 to 3 km of denudation during the Paleocene. In

contrast, the offshore Cenozoic stratigraphic record west of Ireland is mostly intact, with major tectonic activity mostly comprising rifting, and Cenozoic exhumation and deformation [Cook, 1987; Naylor and Anstey, 1987; Roberts, 1989; Shannon, 1991; Naylor, 1992; O'Driscoll *et al.*, 1995; Shannon *et al.*, 1995; Vogt, 1995; Stoker, 1997]. Allen *et al.* [2002] showed that the total volumetric discharge of sediment in the Cenozoic is an order of magnitude smaller than the preserved solid volume of Cenozoic sediment in the basins offshore western Ireland. Therefore the Cenozoic evolution of Ireland must be seen against the broader development of the NW European continental margin.

[6] In this contribution, we show evidence for potential Cenozoic tectonic activity in SE Ireland, which is consistent with observations from other parts of the NW European margin. We first identify tectonic activity on both previously unrecognized faults and reactivated Caledonian and Variscan structures by examining a range of geomorphic indicators of tectonics. We use offshore bathymetric and geophysical data to relate these structures to the inverted Mesozoic Kish Bank Basin off the east coast of Ireland. We then describe onshore evidence of brittle deformation from adjacent areas of the northern margin of the Leinster Granite (Figure 1) that allows us to partially constrain the kinematics of the deformation. Finally, we discuss our structural observations with regards to their likely age and potential geodynamic mechanisms driving this deformation.

2. Regional Setting

[7] The relationship between the topography and surface geology outcrop pattern of Ireland shows oldest rocks forming areas of highest relief and elevation (Figure 1). The only exception to this pattern being the Paleocene-Eocene granite of the Mourne Mountains (Co. Louth and Co. Down) (Figure 1). In many cases, rocks of uniform lithology underlie a wide range of local relief, and therefore areas of relative high relief cannot be explained solely by differential erosion (e.g., enhanced denudation rates of softer rock).

[8] The Irish landscape has been extensively modified by Quaternary glaciers leading to incised U-shaped valleys, corries and glacial depositional features [McCabe, 1996]. However, Mitchell [1980] described evidence for the preservation of extensive preglacial weathering features and suggested that much of the present-day landscape is the product of tectonism, modified by glaciers. Since the end of glaciation in Ireland (circa 10 ka), there has been a relative rise of base level, leading to the drowning of drumlins, eskers and moraines [Gray and Coxon, 1991; Coxon, 2001a].

[9] The present form of the lower Paleozoic inliers and landforms of central and southern Ireland are most likely due to differential erosion between the softer, chemically weathered Carboniferous limestone surface and the more resistant metasedimentary Silurian-Devonian cores [Dewey, 2000; Drew and Jones, 2000; Taylor, 2002; Simms, 2004]

(Figure 1). However, with the notable exception of these inliers, the main areas of relatively high relief tend to occupy the periphery of the island.

[10] The central portion of the Irish landmass has very low relief. Allen *et al.* [2002] used apatite fission track modeling to infer that this area experienced little denudation, with slight local subsidence, during much of the Cenozoic. In general, the area is underlain by Lower Carboniferous carbonate rocks and is cut by a major erosion surface. The low-relief surface extends across part of the Leinster Granite of southeast Ireland [Cunningham *et al.*, 2003] (Figure 1). The presence of nonmarine Oligocene-Miocene clays at Loughshinney [Drew and Jones, 2000], Miocene-lower Pliocene nonmarine clays at Hollymount [Boulter, 1980; Watts, 1985], Pliocene clays at Tynagh [Monaghan and Scannell, 1991], and Pliocene Aeolian sands, lignites and fossilized paleokarstic landforms at Pollnahalia [Coxon and Coxon, 1997] suggests that much of the low-relief land surface is no older than Oligocene (Figure 1).

3. Tectonic Geomorphology

[11] SE Ireland contains the largest contiguous area of high topography on the island (Figure 1). The region is dominated by the Wicklow and Blackstairs Mountains. Topographically, the mountains form a broad, gently rolling upland. The eastern side displays a stepped topography dissected in places by glacially incised valleys up to 200 m deep. The western side slopes more uniformly westward. The topographic massifs of the Wicklow and Blackstairs Mountains do not coincide with the margins of the Leinster Granite (Figures 1 and 2). Plutons of the Leinster Granite are continuous beneath the low-relief Tullow Lowland and are capped with isolated roof pendants of Ordovician schist [Charlesworth, 1963; Brück and O'Connor, 1980] at both high and low elevations. Cunningham *et al.* [2003] used apatite fission track data and geomorphological analysis to infer that the Tullow Lowland has been downthrown to the south by about 700 m in post-early Eocene time. Therefore the differences in elevation across SE Ireland cannot be due simply to differential erosion [Farrington, 1927; Davies, 1966].

[12] The topography of a region (e.g., spatial patterns of elevation and relief, spatial variations in catchment size and relief, map-view patterns of rivers, and river longitudinal profiles) reflects the time-integrated effects of both tectonic and climatic activity. The eastern side of the Wicklow Mountains is marked by a stepped topography dominated by two relatively planar surfaces that gently dip westward, here termed the Djouce and Vartry surfaces (Figure 3).

3.1. Djouce Surface

[13] The Djouce surface forms a relatively high plateau mostly underlain by Leinster Granite (Figure 3). The surface has a modal elevation of 450 m with highest elevations in the east. The bedrock surface dips gently to the west and is incised along the Lough Dan–Tay Fault,

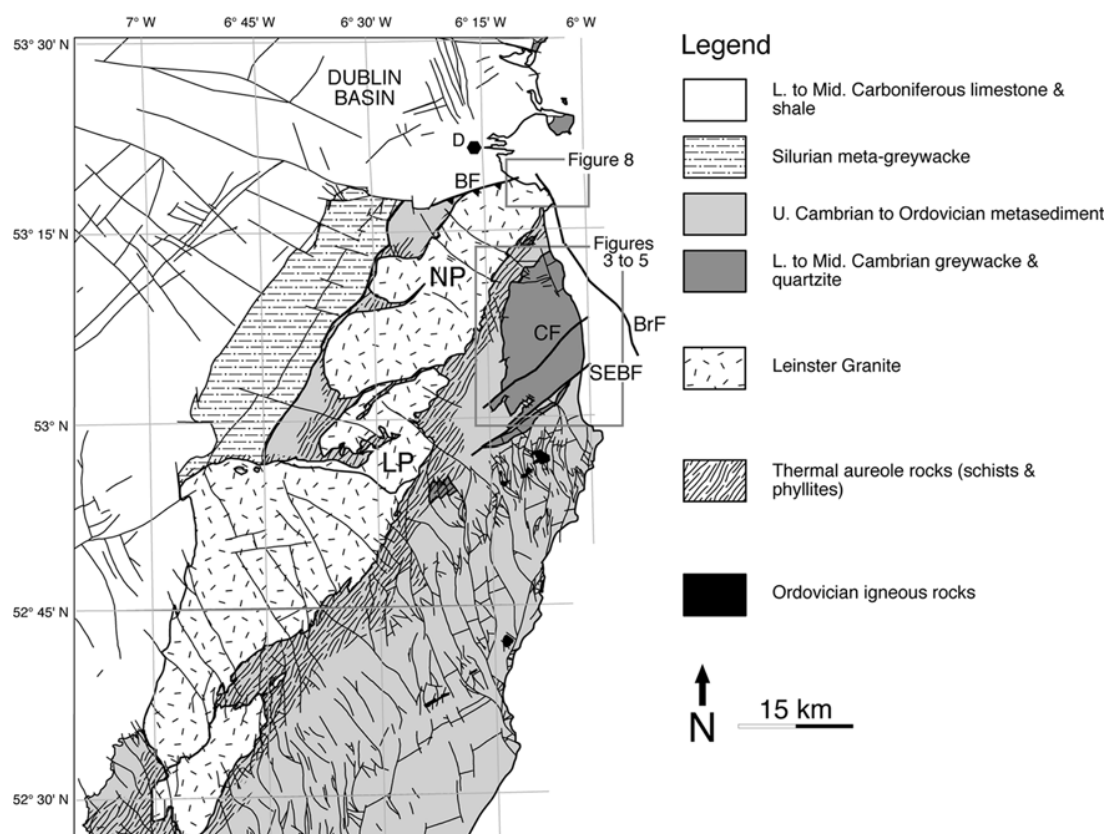


Figure 2. Simplified geological map of SE Ireland [after McConnell and Philcox, 1994; Tietzsch-Tyler and Sleeman, 1994] showing main tectonostratigraphic features. Abbreviations area as follows: BF, Blackrock Fault; BrF, Bray Fault; CF, Callowhill Fault; D, Dublin City; LP, Lugnaquilla Pluton; NP, Northern Pluton; SEBF, Southeastern Boundary Fault.

which separates the Lugnaquilla and Northern plutons (Figure 2). Cunningham *et al.* [2003] inferred a post-early Eocene displacement on the Lough Dan-Tay Fault of 250 m, downthrown to the east. Along the NW trending portion of the fault there is an apparent right-lateral separation of ~500 m on the eastern margin of the granite.

3.2. Vartry Surface

[14] The Vartry surface varies in width from 4 to 10 km, has a modal elevation of 230 m and extends some 20 km from the Dargle River in the north to the Avonmore River in the south (Figure 3a). The surface slopes toward the west, away from the coast, at 1°–2°. Many of the bedrock outcrops exposed on the Vartry surface show deep red staining, alteration of bedrock along joints to red and yellow clay and disaggregation of the underlying greywacke and quartzite. These features provide evidence for paleosol formation under prolonged periods of warm and wet climate [Mitchell, 1980; Wilkinson *et al.*, 1980; Mitchell and Ryan, 1997; Coxon, 2001b]. The lack of compaction of this material and of

the red breccias show that it has never been buried to any great depth.

3.3. Topographic Escarpments

[15] The Djouce surface is separated from the lower Vartry surface by an abrupt, steep linear escarpment with 200 to 250 m relief, here informally named the West Vartry escarpment. The escarpment trend is oblique to the strike of the underlying Cambrian-Lower Ordovician phyllites and metamorphic aureole rocks of the Ribband Group (Figure 3a). Along the foot of the escarpment there are a number of outcrops of very coarse red breccias with a clay matrix and angular boulders of vein quartz and phyllite up to 1 m in diameter (Figure 3a). Likewise, the Vartry surface is bounded on its eastern margin by a second escarpment, here termed the East Vartry escarpment, that steps down to the east with ~100 to 300 m of relief. Again, the escarpment trend is oblique to the strike of the underlying Cambrian greywackes (Figure 3a).

[16] The East and West Vartry escarpments cut across both lithological contacts and the Caledonian structural grain (Figure 3a). Variations in hillslope gradient across

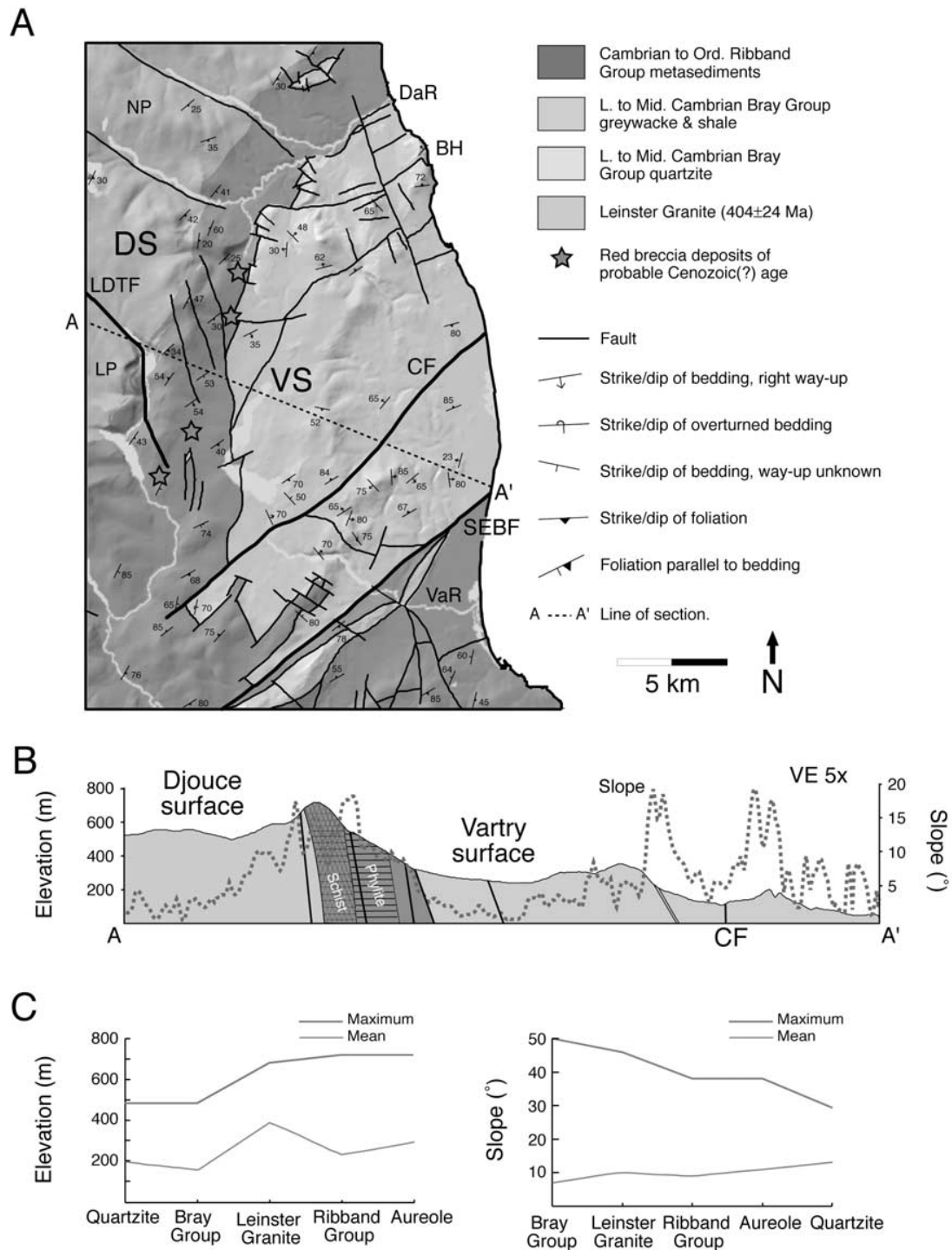


Figure 3. (a) Bedrock geology shaded relief image of SE Ireland. Abbreviations area as follows: BH, Bray Head; CF, Callowhill Fault; DaR, River Dargle; DS, Djouce surface; LDTF, Lough Dan-Tay Fault; SEBF, Southeastern Boundary Fault; VaR, Vartry River; VS, Vartry surface. (b) WNW-ESE trending cross section showing elevation, topographic slope (dashed line, calculated from the digital elevation model using 3×3 window of 50 m cells), and bedrock geology. (c) Mean and maximum values of elevation and slope for the various bedrock lithologies. Aureole rocks and quartzite are included within Ribband and Bray Groups, respectively, but are also shown separately as a guide to their effects on slope. See color version of this figure at back of this issue.

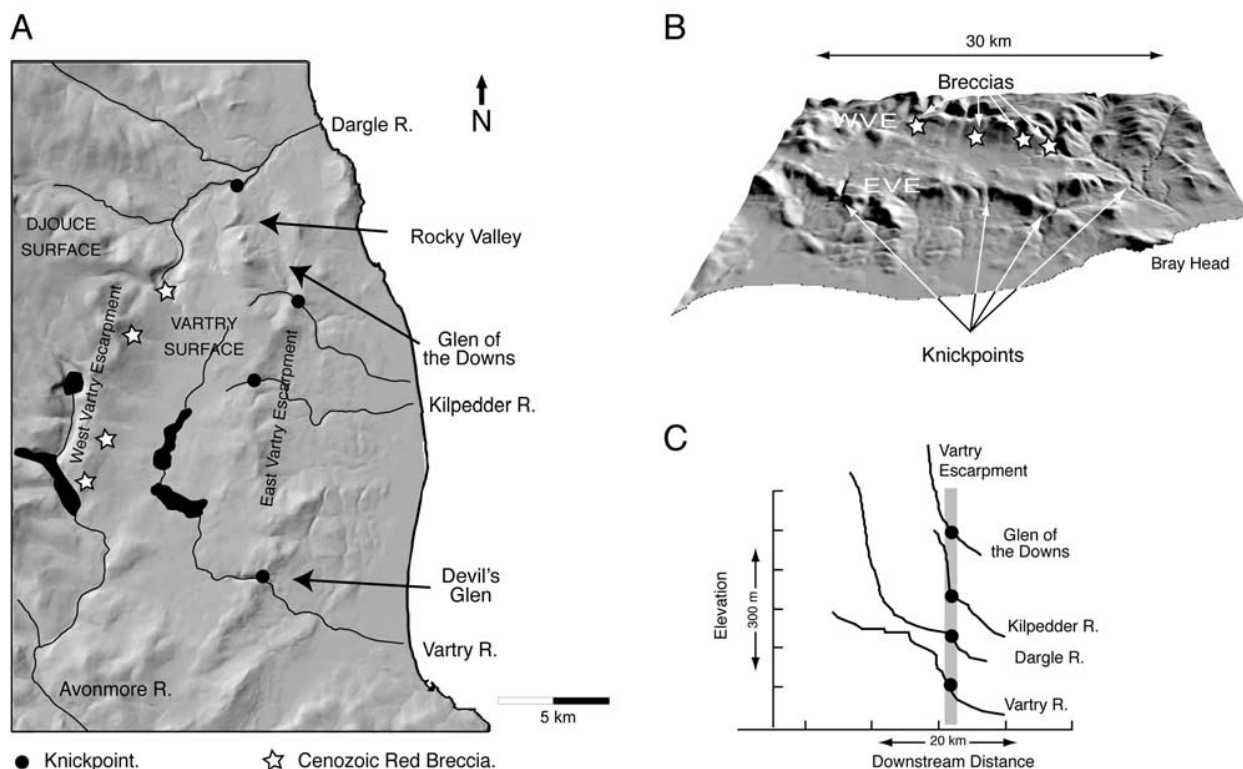


Figure 4. (a) Shaded relief image of SE Ireland showing major rivers, the Djouce and Vartry surfaces, and the East and West Vartry escarpments. Stars indicate exposures of red breccias described in text. Solid dots show the locations of knickpoints identified in Figure 4c. (b) Three-dimensional view of the region, illustrating the East and West Vartry escarpments (EVE and WVE, respectively). (c) Longitudinal profiles of streams flowing across the East Vartry escarpment, which is denoted by gray bar.

the escarpments do not always correspond with changes in lithology (Figure 3b). The relatively erodible greywackes form areas of low relief and elevation, while aureole rocks and quartzites generally produce steep slopes and areas of relatively high relief (Figure 3c). This implies that landscape evolution is partly controlled by differential erosion. However, the greywackes of the Bray Group atypically underlie the steepest slopes in the Vartry area, e.g., at Bray Head and along the East Vartry escarpment (Figures 3b and 3c). It is therefore unlikely that the escarpments are a product of resistant bedrock. The steepness of the walls implies that the hillslope gradients are being actively maintained by fluvial incision in response to base level fall [e.g., *Densmore et al.*, 1997].

3.4. Fluvial Geomorphology

[17] The Vartry surface is drained by (1) a series of small catchments that flow to the east, across the East Vartry escarpment and (2) the Vartry River, which flows south along the long axis of the surface before turning abruptly east (Figure 4). Each of these streams has incised a steep-walled, bedrock gorge across the East Vartry escarpment; the depth of each gorge is partly related to the drainage area of the catchment, and varies from 25 m for the smallest

catchments to 120 m for the Vartry River, which forms the spectacular Devil's Glen (Figure 4). Longitudinal profiles of the rivers that cross the East Vartry escarpment show bedrock knickpoints of between 25 and 120 m, consistent with a fall in base level relative to the headwaters of each catchment.

3.5. Interpretation

[18] As noted above, the escarpments are unlikely to be due to differential erosion. The elevation of bedrock at the base of the East Vartry escarpment varies from 20 to 140 m, which is inconsistent with a possible origin as a sea cliff. Marine or glacial erosion processes of the East Vartry escarpment do not explain the westward tilt of the Vartry surface, nor its sharp southern termination, just south of the Devil's Glen (Figure 4). *Davies* [1966] argued that the Vartry surface was a series of subaerial pediments cut into bedrock, but this explanation fails to explain the limited spatial extent of the surface, its westward tilt, and the abrupt escarpments at its eastern and western margins. Therefore it is likely that the knickpoints on the East Vartry escarpment are due to faulting. The occurrence of red breccias along the West Vartry escarpment likewise implies faulting. Hence we interpret the escarpments as the

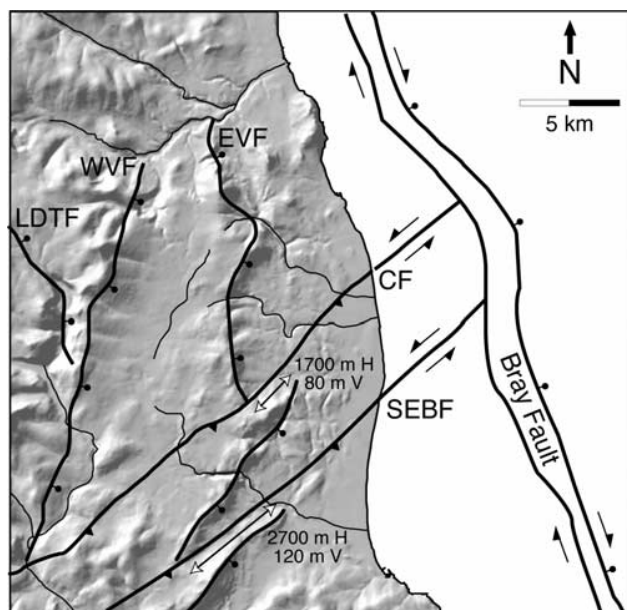


Figure 5. Shaded-relief image of SE Ireland, showing locations and inferred kinematics of Cenozoic faults (East and West Vartry faults) and apatite fission track data (LDTF is Lough Dan-Tay Fault of *Cunningham et al.* [2003]). Note the close spatial relationship between these structures and the offshore dextral-normal Bray Fault as mapped from seismic and aeromagnetic data. In this model, NNW striking faults, parallel to the Bray Fault, are assumed to be dextral transtensional structures, whereas those faults oriented N–S (e.g., the East and West Vartry faults) are primarily extensional. NE striking faults have sinistral transpressional kinematics; this is consistent with topographic offsets of the East Vartry escarpment across the Callowhill and Southeastern Boundary Faults (H, horizontal offset; V, vertical offset). Balls on normal fault hanging walls, barbs on thrust fault hanging walls. Abbreviations are as follows: EVF, East Vartry fault; LDTF, Lough Dan-Tay Fault; WVF, West Vartry fault.

topographic expression of faults. These structures are subparallel to the Bray Fault that lies between 1 and 5 km offshore to the east (Figure 5).

[19] South of the Devil's Glen, the East Vartry escarpment is segmented en echelon and changes strike from NE to ENE. The change in strike corresponds with two known Caledonian faults that appear to have offset the escarpment. The Callowhill fault sinistally offsets the escarpment by ~ 1.7 km (as measured by offset of the 200 m contour), and the maximum topographic elevation is 80 m higher on the south side of the fault. Sinistral offset of the escarpment across the Southeastern Boundary fault is ~ 2.7 km, and the maximum elevation is 120 m higher on the south side (Figure 5). This implies almost identical sinistral transpressional kinematics on both structures, each with a strike-slip to dip-slip ratio of $\sim 20:1$.

[20] These relationships imply that dip-slip motion occurred along the eastern margin of the Wicklow Mountains, displacement being taken up by faults that form the West and East Vartry escarpments and that are subparallel to the offshore Bray Fault. During or after this deformation, sinistral transpressional displacement occurred along NE trending, reactivated Caledonian structures (Figure 5). Below, we describe geophysical evidence for activity along the offshore continuation of these structures in the Kish Bank Basin.

4. Offshore Structures in the Kish Bank Basin, Irish Sea

[21] To the east of the Wicklow and Blackstairs Mountains, between 2 and 10 km offshore (Figure 6a), Upper Carboniferous and Mesozoic sedimentary rocks are preserved in a down-faulted outlier, the Kish Bank Basin. The basin is thought to represent the erosional remnant of a larger, linked Mesozoic basin system that covered much of the Irish Sea and western Britain [e.g., *Dunford et al.*, 2001] and probably extended into the north of Ireland as shown by onshore successions near the Kingscourt Outlier [*Visscher*, 1971; *Naylor et al.*, 1993] and the Ulster Basin, Co. Antrim [*Manning and Wilson*, 1975]. Within the basin, a thick Permo-Triassic package unconformably overlies Westphalian rocks [e.g., *Jenner*, 1981; *Naylor et al.*, 1993; *Corcoran and Clayton*, 2001; *Dunford et al.*, 2001]. The bulk of the extension occurred during Permo-Triassic times [*Shannon*, 1991; *Naylor et al.*, 1993; *Dunford et al.*, 2001]. Interpretation of vitrinite reflectance profiles (Shell 33/22-1) from Upper Carboniferous sequences (Figure 6a) indicates that between ~ 1.5 km [*Corcoran and Clayton*, 2001] and 3 km [*Jenner*, 1981] of section has been removed.

[22] The basin is divided in two parts by the NNW trending Codling Fault. The fault zone is 2 to 5 km wide and dextrally offsets the Dalkey and Lambay faults by ~ 4 km [*Broughan et al.*, 1989] (Figures 6a and 6c). Recent seismic studies suggest that as much as 9 km of strike-slip displacement has occurred during the Cenozoic [*Dunford et al.*, 2001]. Thickening of Neogene-Quaternary sediments is observable on seismic profiles and 381 m were encountered in well Fina 33/17-1. The base of the Neogene-Quaternary section has been tentatively dated as lower Pliocene from poor miospore recoveries [*Naylor et al.*, 1993]. On the basis of similar orientation and sense of displacement, a number of previous workers have suggested that the Codling Fault is aligned with the dextral-normal Newry Fault in NE Ireland (see Figure 9 for location), which is known to have been active in syn- and post-Oligocene time [*Charlesworth*, 1963; *George*, 1967; *Jenner*, 1981; *Kerr*, 1987; *Parnell et al.*, 1988; *Readman et al.*, 1997; *Geoffroy et al.*, 1996].

[23] The NNW trending Bray Fault is parallel to the Codling Fault, and lies ~ 1 km offshore from Bray Head. This fault brings Cambrian rocks against Lias. *Broughan et al.* [1989] suggested that as much as 2.7 km of Liassic and younger section may be preserved in the hanging wall of the Bray Fault (Figure 6a). On the basis of known thicknesses of missing Mesozoic and lower to upper Paleozoic strati-

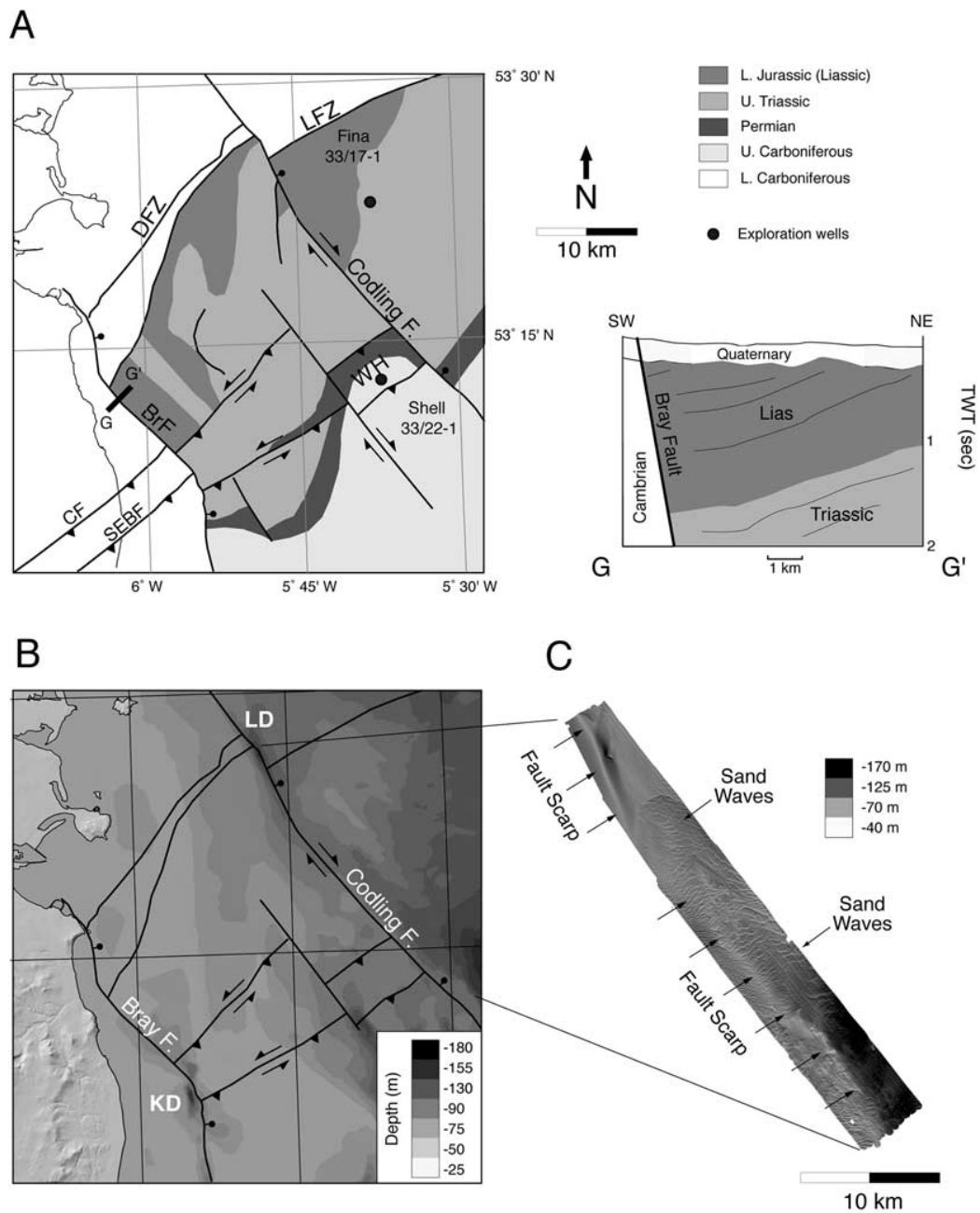


Figure 6. (a) Pre-Cenozoic geological map and structural framework of the Kish Bank Basin (modified from Jenner [1981]). Seismic reflection profile (marked by G–G' on map) shows postulated 2.7 km of preserved Lias section forming hanging wall rocks of the Bray Fault (BrF) [Broughan *et al.*, 1989]. Abbreviations are as follows: BrF, Bray Fault; CF, Callowhill Fault; DFZ, Dalkey Fault Zone; LFZ, Lambay Fault Zone; SEBF, Southeastern Boundary Fault; WH, Wickow High. (b) Bathymetry of the western Irish Sea showing the east facing bathymetric scarp of the Codling Fault. Where the Codling and Bray Faults trend in a northerly direction, small bathymetric deeps with relief of ~80 m have formed, e.g., the Kilcoole Deep (KD) and the Lambay Deep (LD). (c) Multibeam bathymetric shaded relief showing relief along Codling Fault. Note the larger wavelength of sand waves to the east of the fault scarp, indicating deeper water.

graphic successions from both the Kish Bank Basin and onshore Dublin Basin, this gives an inferred minimum throw of 4 km. Interpretation of seismic reflection profiles and aeromagnetic data shows that the postulated Lias section pinches out to the east and is faulted (northern margin of Wicklow High block) against Triassic rocks to the southeast (Figure 6a).

4.1. Bathymetry

[24] Water depths just west of the Codling Fault are on average 50 to 70 m (Figure 6b). To the east, the fault zone forms an east facing scarp on the seabed, with a maximum bathymetric relief of ~80 m (Figures 6b and 6c). Recent multibeam bathymetric surveys have imaged the surface morphology of the fault zone and show strong variation in scarp relief along the structure (Figure 6c). There is no seismic reflection evidence of Triassic synrift sedimentation [e.g., *Naylor et al.*, 1993]. Hence the fault zone was not active during Triassic rifting. Dextral transtensional displacements are therefore Cenozoic. This is supported by the presence of active gas seepage along the fault trace [*Corcoran and Clayton*, 1999] and microseismic activity in 1982. Toward the northern end of the Codling Fault, water depths attain a maximum of 150 m within the bathymetric Lambay Deep (Figures 6b and 6c). The deep is oriented NNW-SSE and is <1 km by <0.5 km in dimension. A similar, NNW trending bathymetric feature occurs ~10 km east of the Vartry surface along the Bray Fault (Figure 6b). Water depths in the deep reach a maximum of 120 m in contrast with ~20 to 40 m on adjacent shelf areas. Both bathymetric deeps have length to width ratios of ~3:1 and are located where the Codling and Bray Faults swing to a more northerly strike (Figure 6b). We interpret their formation as the product of dextral transtension, producing localized extension where the fault strands are in a favorable northerly strike.

4.2. High-Resolution Aeromagnetic Data

[25] The shallow structure of the Kish Bank Basin is well imaged by a pseudodepth slice filtered high-resolution magnetic intensity map (Figure 7). The magnetic image shows thick basin fill and deep basement as regions of relatively low magnetic intensities. At the southern margin of the basin, magnetic intensities are relatively high and thus depict areas of shallow basement. This was proven by Shell 33/22-1 well that encountered a thin Neogene-Quaternary section (71 m) overlying 720 m of Upper Carboniferous (Westphalian) succession, unconformably resting on basement of lower Paleozoic slate. The lower Paleozoic slates have previously been correlated with the near onshore Bray Group (Figure 6a) [*Jenner*, 1981; *Naylor et al.*, 1993].

[26] High magnetic intensity anomalies correspond with the known Ordovician basic volcanic rocks on Lambay Island (LI on Figure 7). A negative anomaly (PI on Figure 7) occurs just east of the Codling Fault, about 6 km south of the Lambay Deep. The high intensity of the anomaly suggests it is probably an igneous intrusion. The

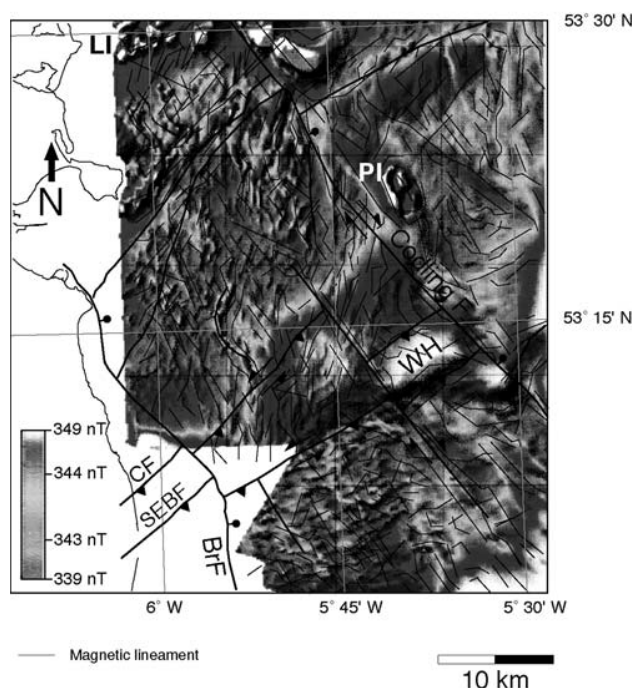


Figure 7. Total magnetic intensity image of the Kish Bank Basin. Image area is the same as in Figures 6a and 6b, and fault traces are overlain for comparison. Magnetic intensity is reduced to pole with a pseudo depth slice filter applied. The filter accentuates high-frequency, low-amplitude shallow magnetic structure. Note trends in major magnetic lineaments indicated by thin lines (see section 4.2 for full discussion). Abbreviations are as follows: BrF, Bray Fault; CF, Callowhill Fault; LI, Lambay Island; PI, Intrusion; SEBF, Southeastern Boundary Fault; WH, Wicklow High. See color version of this figure at back of this issue.

anomaly is elliptical in plan view with the long axis oriented parallel to the Codling Fault. The axis swings from a NNW to NNE orientation at the northern end of the body. The clockwise swing, subtle S-shape geometry and parallelism of the axis with the Codling Fault implies that the intrusion is syntectonic. As outlined in section 1, there was much igneous activity during the Paleogene, and the Codling Fault was active at this time [e.g., *Charlesworth*, 1963; *Naylor et al.*, 1993]. The intrusion displays reverse polarity, which is typical of Paleogene dikes, sills and lava flows in Ireland and Britain [e.g., *Saunders et al.*, 1997]. Thus we infer that the anomaly is most likely a Paleogene igneous intrusion related to the North Atlantic Igneous Tertiary Province.

[27] Directional filtering of the pseudodepth sliced magnetic image enhances relatively shallow magnetic lineaments (Figure 7). The Codling, Lambay, and Dalkey faults are delineated by subtle changes in lineament pattern and by sharp, discontinuous linear anomalies above the fault zones. Shallow anomalies north of the axis of the Wicklow High basement ridge are dominantly

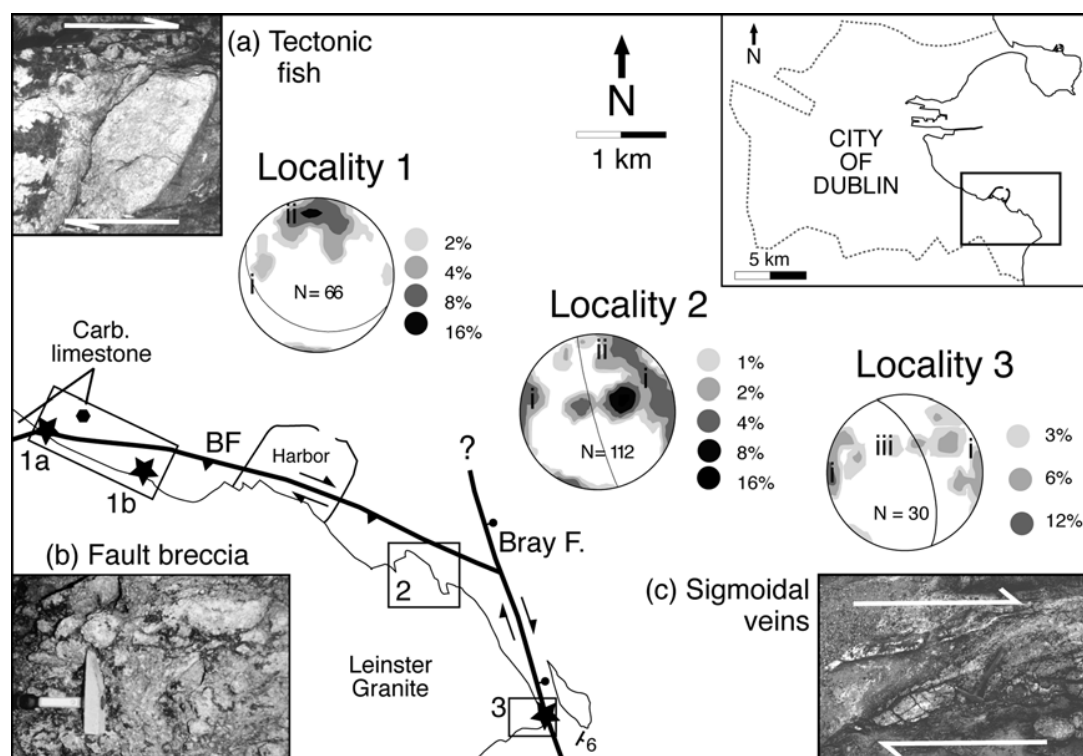


Figure 8. Geological map and structural data from the northern margin of the Leinster Granite. Poles to structural shear planes and foliation plotted on equal area, lower hemisphere stereonets; density contour interval 1%. Locality 1 is dominated by WNW-ESE cataclastic semibrittle structures (shown as “ii” on the stereonets); sense of shear is dextral-reverse. Locality 2 is dominated by NNW-SSE structures with sinistral sense of shear (shown as “i” on the stereonets). Locality 3 is dominated by NNW-SSE sinistral-normal structures, cut by NW-SE dextral-normal shear zones (shown as “iii” on the stereonets) that form z-shaped sigmoidal lineations plunging gently to the north. Solid stars are locations of fault breccias/cataclasite. Ticks on downthrown block. Note “mean plane” annotation on stereonets for major structural groups. BF is Blackrock Fault. Insets are as follows: (a) tectonic fish at locality 1a showing a dextral-reverse sense of shear. View toward the SSE. (b) Fault breccia at locality 1a, where the Blackrock Fault swings from an ENE to a WNW trend. View toward the NNE. (c) Dextral sigmoidal deformation of pegmatite vein within NW-SE brittle dextral-normal shear zone at locality 3. View toward the SW.

north or NE trending. South of the ridge axis, the dominant lineament trend changes to ENE, indicating the potential presence of a significant structural break associated with the southern margin of the Wicklow High (Figure 7). Small z-shaped sigmoidal changes of magnetic lineament are observable along the southern margin of the High and may suggest the presence of a fault with a component of left-lateral displacement. This sense of displacement is consistent with the same Cenozoic stress field that has driven slip on both the Codling and Newry fault systems. On the basis of this evidence, we tentatively interpret the fault bounding the southern and northern margin of the Wicklow High as being coupled with the dextral-normal Codling Fault. The intersection angle of the faults is $>60^\circ$, suggesting sinistral transpression. When traced onshore, the faults correlate with

the onshore Callowhill and Southeastern Boundary faults (Figure 7).

5. Brittle-Ductile Structures in the Northern Leinster Granite

[28] A corollary of near-offshore Cenozoic faults is the potential for tracing such structures onshore and the potential for reactivation of preexisting shear zones. To test this hypothesis, we mapped structures cutting onshore outcrops along the northern margin of the Leinster Granite.

[29] *Brindley* [1954] recognized three important phases of deformation along the northern margin of the Leinster Granite, categorized as (1) primary flow textures; (2) primary fracture features (e.g., pegmatite-aplite dike sys-

tems); and (3) later mechanical features (e.g., jointing, cataclasis). The “later mechanical features” were previously described as postconsolidation effects in the form of marginal zones of shearing [Brindley, 1954] and include both brittle fracture systems and ductile textures. On steeply inclined foliation surfaces, Brindley [1954] observed intergranular deformation associated with the development of secondary muscovite along the surfaces.

[30] In this study, detailed remapping along the northern margin of the Leinster Granite reveals three sets of post-Caledonian structures associated with shear zones, all part of Brindley’s “later mechanical features,” that postdate granite emplacement. These structures are distinguished on the basis of orientation, kinematics, deformational style and overprinting relationships (Figure 8).

5.1. Brittle-Ductile NNW-SSE Sinistral Shear Zones

[31] On the northern margin of the Leinster Granite, synmagmatic fabrics are offset by NNW-SSE striking, near-vertical deformation zones (group “i” data on Figure 8). These structures are observed around the entire northern margin of the batholith and are particularly prominent at localities 2 and 3 (Figure 8). The shear zones form 1–5 m wide bands spaced about 30 m apart, and display a well-developed, moderately east dipping foliation. Kinematic indicators on the margins of the shear zones predominantly show a sinistral strike-slip sense of shear and include asymmetrical S-C fabrics and offset pegmatite dikes. In places, slickensides and striae pitch steeply to the south on near-vertical planes, consistent with up to the east sense of shear. The shear zones at localities 2 and 3 are geometrically and kinematically identical. Therefore we correlate these structures as part of the same shear zone.

5.2. Semibrittle WSW-ENE to WNW-ESE Dextral-Reverse Shear Zones

[32] The WSW-ENE trending Blackrock Fault is parallel to the Caledonian fabric on the NW margin of the Leinster Granite and lower Paleozoic aureole rocks. The fault swings to a WNW-ESE trend where it reaches the coastline, and dextral-reverse shear zones are intensely developed adjacent to the fault (group [ii] data on Figure 8). The deformation is brittle in nature, with the Leinster Granite thrust northward over the Viséan limestones of the Dublin Basin.

[33] The shear zones are observed in coastal sections and have a consistent WNW-ESE strike at localities 1 and 2 (Figure 8). The main fault zone lies offshore with smaller splays on the shear zone margin exposed onshore. In plan view, the WNW-ESE shear zones form an anastomosing network enclosing lenticular lozenges of less-deformed granite. Since much of the outcrop is restricted to coastal sections, the shear zone margins are rarely exposed. These structures offset the NNW-SSE sinistral shear zones described in section 5.1.

[34] The mesoscopic structures that comprise these shear zones of the Blackrock Fault typify semibrittle deformation. We interpret the bulk deformation of the shear zones as a progressive dextral shear based on the following observations. First, sigmoidal “fish,” with length to width ratios of

about 2:1, possess an S-shape geometry and are exposed on semivertical surfaces, showing consistent dextral-reverse (top to the north) kinematics (Figure 8). Second, sigmoidal foliation has a mean strike of $\sim 090^\circ$ and the finite plane or line of transport trends $\sim 110^\circ$. This reflects a progressive deformation where increments of infinitesimal strain produce C-planes that are subsequently deformed, displaying consistent dextral kinematics. Finally, a lozenge-shaped band of cataclasite, up to 2 m thick and striking WNW-ESE, is exposed near Seapoint (locality 1b, Figure 7). Relict granitic clasts within a fine-grained matrix show crude sigmoidal S- and Z-geometries on a subhorizontal surface.

[35] A breccia is exposed about 300 m to the west at locality 1a (Figure 8), on the Blackrock Fault contact between the Leinster Granite and Carboniferous limestones. The fault breccia is notably angular, with pieces of granite (up to 30 cm in diameter) and occasional limestone and shale clasts set in a brecciated groundmass. The clasts are aligned on an E-W striking pressure solution cleavage and their long axes plunge at about 20° to the east. The cleavage steepens from a dip of 50° SW to 80° NE and swings clockwise in strike toward the fault. This indicates a predominantly dextral shear on the Blackrock Fault with a component of upthrow on the south side.

5.3. NW-SE Brittle Dextral-Normal Shear Zones

[36] A younger set of shear zones cut the NNW-SSE sinistral shear bands at localities 2 and 3 (group [iii] structures on Figure 8) and offset the WNW-ESE structures at locality 2. These structures trend NW-SE and dip between 60° and 84° to the NE. Individual zones range from a few meters to more than 100 m in width, and can be traced for tens of meters along strike. Two zones at locality 3 (Figure 8) contain S-shaped sigmoidal quartz veins and cataclasite with rotated porphyroclasts, indicating dextral shear. A stretching lineation on the shear zone walls at locality 3 plunges moderately to the SE and is consistent with dextral-normal shear (down to the east). We speculate that these structures are the onshore continuation of the Bray Fault (see section 6).

5.4. Evidence of Timing for Onshore Structures

[37] The WNW-ESE dextral and NNW-SSE sinistral shear zones are compatible with a NW directed maximum horizontal compression. A pressure solution cleavage is confined to the NNW-SSE sinistral shear zones and deformational style is brittle-ductile. It is possible that the NNW-SSE sinistral shear zones are reidal shears to the WNW-ESE dextral semi brittle structures. However, Late Carboniferous to Early Permian Variscan deformation [Graham, 2001] of the Viséan rocks of the Dublin Basin consists of E-W trending fold arrays in the northern part of the basin, related to NW-SE dextral shear zones and N-S trending sinistral shear zones [Johnston, 1993]. The WNW-ESE dextral-reverse structures, on the northern margin of the granite, offset the NNW-SSE sinistral shear zones (Figure 8). Therefore it is likely that the NNW-SSE sinistral shear zones are the product of Variscan deformation.

[38] The WNW-ESE dextral-reverse shear zones of the Blackrock Fault are notable for their brittle cataclastic textures where, at the coast, the fault changes strike from ENE-WSW to WNW-ESE. In general, large quartz grains display undulose extinction with partial polygonization. Secondary growth of muscovite has facilitated cataclastic flow, as the micas have been smeared out along the planes of the original magmatic foliation. The breccia at locality 1a lies along the Blackrock Fault and contains angular blocks of granite set in a fine-grained cataclastic matrix with rare clasts of Viséan shale and limestone. The Viséan clasts show that the breccia is not of Caledonian origin [Brindley, 1957] and can be no older than Variscan in age.

[39] Vitrinite reflectance data indicate that, during Variscan deformation (Upper Carboniferous to Lower Permian), Viséan limestones cropping out in the Dublin region reached peak paleotemperatures of $\sim 350^{\circ}\text{C}$ in a geothermal gradient of $\sim 50^{\circ}\text{C km}^{-1}$ [Clayton *et al.*, 1989]. The known thickness of limestone in the contemporary Dublin Basin is about 2.5 km. The granite outcrops currently exposed along the northern margin of the batholith lay about 0.5 km below the Carboniferous unconformity [e.g., Graham, 2001], and therefore must have attained temperatures of at least 400°C during Variscan deformation. As the brittle-ductile transition for quartz occurs at about 300°C [Kerrick *et al.*, 1977; Barker, 1998], the presence of unpolygonized quartz grains argues that the WNW-ESE dextral-reverse shear zones represent post-Variscan deformation.

[40] The Blackrock Fault, in places, forms a topographic escarpment between the lower Paleozoic inlier and the postulated Oligocene erosion surface (see section 2 above). Therefore, while not conclusive, it may be tentatively inferred from this evidence that reactivation of the Blackrock Fault is of at least Oligocene age as is the erosional surface.

[41] The youngest structures, on the NE margin of the batholith, are NW trending shear zones that offset both earlier sets of structures. The kinematics of the structures are consistent with dextral-normal (down to the east) displacement, and lie approximately along strike from the offshore Bray Fault (Figure 8).

6. Discussion

6.1. Geomorphological Constraints

[42] The presence of knickpoints on the rivers draining the Vartry surface implies that insufficient time has passed since knickpoint generation for the rivers to re-achieve equilibrium long profiles. With a base level fall caused by a drop in sea level or by an increment of tectonic displacement along a river that is in equilibrium with its boundary conditions, a wave of incision will propagate upstream from the locus of base level change. The adjustment timescale t_a reflects the time necessary for this wave to reach the upstream reaches of the catchment. For times $t < t_a$, the river will appear to be out of equilibrium; a knickpoint will separate reaches of the river that have equilibrated (through incision) from those that have not. For $t > t_a$, the river will appear to be in equilibrium with its new boundary con-

ditions. Thus the presence of knickpoints in the catchments draining the Vartry surface argues that the most recent disturbance on those catchments must have occurred less than t_a years ago. In addition, the spatial correspondence between the knickpoints and the East Vartry escarpment implies that the knickpoints have not had enough time to migrate significant distances upstream.

[43] In general, rates of knickpoint migration are not well documented, particularly in relation to faulting. Nevertheless, the presence of knickpoints on the East Vartry escarpment places some constraint on the timing of knickpoint generation. Pazzaglia *et al.* [1998] and Young and McDougall [1993] have independently shown that knickpoints may survive on catchments draining passive margins for ~ 10 Myr. Pazzaglia *et al.* [1998] demonstrated that knickpoints in the modern bed of the Susquehanna River, a large catchment in the eastern United States, were visible in the profile of a fluvial terrace dated to ~ 10 Ma. Young and McDougall [1993] compared modern profiles of small catchments in New South Wales, Australia, with paleoprofiles preserved by a basalt flow, and demonstrated that knickpoints and profile irregularities were not removed in 20 Myr. These results, while not directly applicable to SE Ireland, do suggest that the knickpoints along the East Vartry escarpment may have persisted for ~ 10 –20 Myr, or possibly longer.

6.2. Onshore-Offshore Correlation and Timing

[44] The escarpments that bound the eastern margins of the Djouce and Vartry surfaces are interpreted as the surface expression of Cenozoic faults. Displacements on these faults, which we informally refer to as the East and West Vartry faults, must be relatively recent to explain the existence of knickpoints, the discordance between foliation and topography, the occurrence of reddened fault breccias, and the lack of compaction in these breccias (particularly as apatite fission track modeling predicts large amounts of exhumation during Cenozoic time [Allen *et al.*, 2002]). The East and West Vartry faults are subparallel to the offshore dextral-normal, post-Liassic Bray Fault and post-Oligocene Codling Fault (Figures 5 and 6). Like the offshore structures, the onshore faults have a down to the east sense of vertical displacement. Likewise, the NW-SE dextral-normal shear zones on the NE margin of the Leinster Granite at locality 3 lie approximately along strike of the Bray Fault (Figure 8). If we make the assumptions, on the basis of geometry, kinematics, and occurrence of bathymetric deeps, that (1) the Bray Fault was reactivated by the same stress system which has driven late Cenozoic offsets on the Codling Fault and (2) the Vartry faults and the onshore shear zones at locality 3 are part of the same fault zone as the Bray Fault, then displacement on the Vartry faults and the locality 3 shear zones are at most post-Liassic, and probably post-Oligocene, in age.

[45] The aeromagnetic data show two major faults bounding the offshore Wicklow High block can be extended westward to the onshore Callowhill and Southeastern Boundary faults (Figures 5 and 7; described in section 4.2 above). These faults, in turn, have been inter-

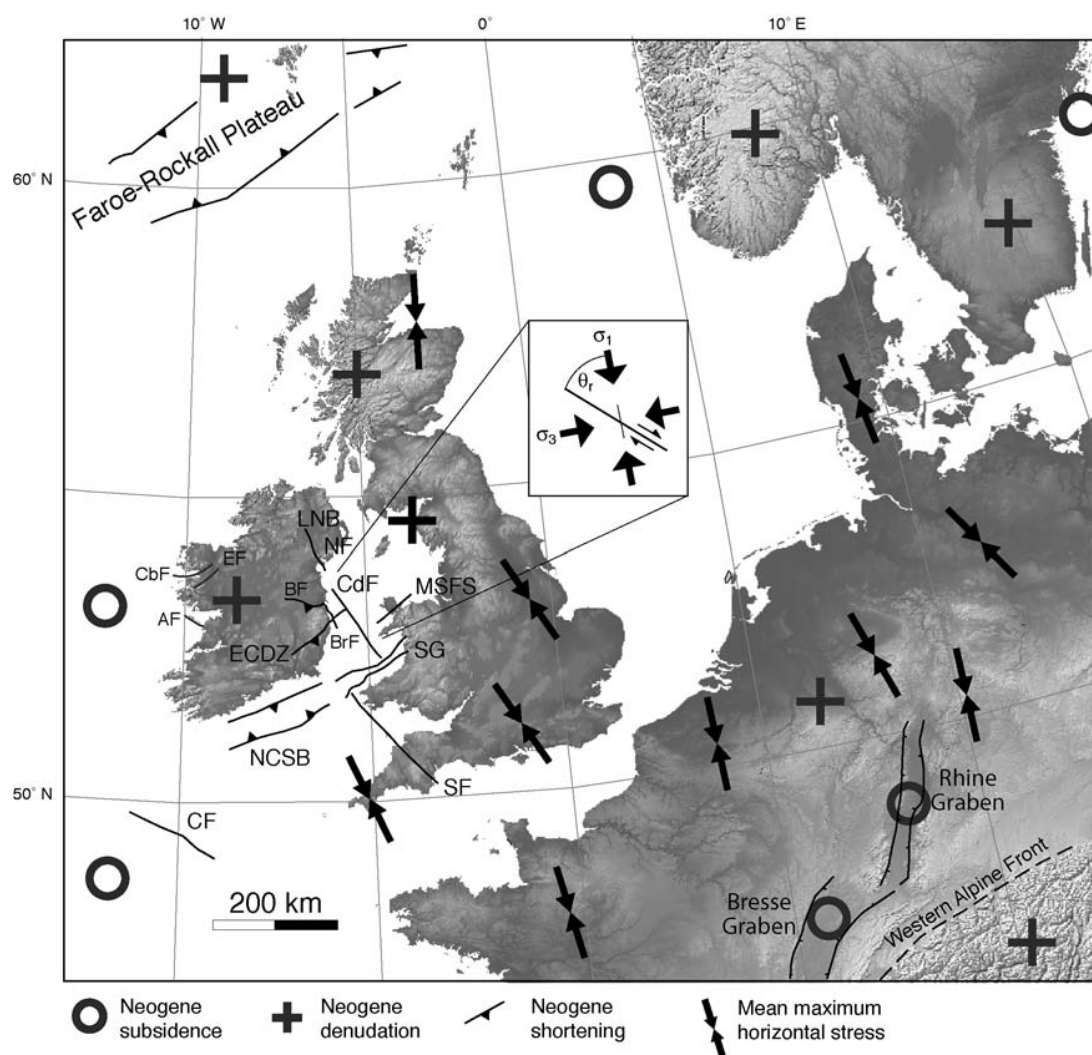


Figure 9. Summary shaded relief image of NW Europe showing Cenozoic structures in Ireland and Britain, extensional graben in the Alpine foreland, and compression on the Faeroe-Rockall Plateau [Coward *et al.*, 1989; Boldreel and Andersen, 1998; Cunningham *et al.*, 2003; Parnell *et al.*, 1988]. Also shown are generalized areas of Neogene uplift and subsidence [after Japsen and Chalmers, 2000]. Thick black arrows show the mean maximum compressive horizontal stress (S_{hmax}) [Müller *et al.*, 1992, 1997]. Inset shows post-early Eocene proposed paleostress directions for SE Ireland (symbols after Sibson [2001]). Abbreviations are as follows: AF, Aran Fault; BF, Blackrock Fault; BrF, Bray Fault; CdF, Codling Fault; CbF, Clew Bay Fault; CF, Cockburn Fault; ECDZ, East Carlow Deformation Zone; EF, Erriff Fault; LNB, Lough Neagh-Ballymoney basins; MSFS, Menai Strait fault system; NCSB, North Celtic Sea Basin; NF, Newry Fault; SF, Sticklepath-Lustleigh Fault; SG, St George's Fault.

preted as the northern extension of the southeast dipping, Caledonian-age, ductile East Carlow Deformation Zone (ECDZ; Figure 9) [McArdle and Kennedy, 1985; McArdle *et al.*, 1987]. Cunningham *et al.* [2003] suggested, on the basis of apatite fission track data and geomorphology, that the ECDZ was reactivated post-early Eocene time. The Callowhill and Southeastern Boundary faults clearly displace the East Vartry fault, and correlation with the ECDZ suggests that this displacement also took place in post-early Eocene time. This timing is further

constrained by (1) the Blackrock Fault offset of the postulated Oligocene surface to the north (see section 2), (2) known post-Oligocene deformation on the Codling Fault, and (3) the occurrence of knickpoints along the East Vartry fault.

6.3. Tectonic Model

[46] We infer that evidence for two separate Cenozoic deformation events are preserved in SE Ireland. The first led to reactivation of the Caledonian–Variscan Blackrock Fault

and is compatible with NW-SE σ_1 , subvertical σ_2 and NE-SW σ_3 . A subsequent clockwise swing of σ_1 to the NNW-SSE is compatible with the following fault kinematics: (1) normal (dextral?) displacement on the West and East Vartry faults, and dextral transtension on the Codling and Bray faults, (2) NW-SE dextral-normal brittle shear zones cutting the NE margin of the Leinster Granite, and (3) sinistral transpression on the fault strands that bound the Wicklow High block. The reasons for the subtle clockwise swing in stress orientations between the two deformation events is unclear, but *Gölke et al.* [1996] showed that local structures can cause significant modifications of the regional stress field.

[47] The youngest stress system is compatible with the syn- and post-Oligocene stress system inferred from the Lough Neagh-Ballymoney basins of NE Ireland [*Geoffroy et al.*, 1996; *Kerr*, 1987; *Mitchell*, 1980; *Parnell et al.*, 1988; *Solla-Hach*, 2002] and the Cenozoic stress system inferred from western Ireland [*Dewey*, 2000] (Figure 9). It is also in harmony with Eocene to mid-Oligocene displacement on the dextral-reverse, NW-SE trending Sticklepath-Lustleigh fault and pull-apart basins of SW England (Figure 9) [*Bristow and Hughes*, 1971; *Blundell*, 2002], sinistral displacement of Cenozoic dike swarms along the NE-SW trending Menai Strait fault system of north Wales [*Bevins et al.*, 1996], early Neogene shortening in the Cockburn Basin [*Smith*, 1995], and SE directed compression from early Eocene time onward across much of the NW European margin [*Doré et al.*, 1999] (Figure 9).

6.4. Mechanisms

[48] Regional apatite fission track studies consistently show evidence for Cenozoic cooling events on the NW European margin. A major one occurred at ~30 Ma with 1 to 3 km of denudation [*Lewis et al.*, 1992; *Rohrman et al.*, 1995; *Green et al.*, 1993, 2000; *Allen et al.*, 2002]. Compaction studies and burial estimates suggest similar magnitudes of denudation [e.g., *Bulat and Stoker*, 1987; *Murdoch et al.*, 1995; *Japsen*, 1997]. *Japsen and Chalmers* [2000] summarized evidence for, and attempted to link, Cenozoic tectonic activity and denudation around the North Atlantic. A number of studies have argued for either rock uplift or denudation, or both, along various parts of the margin. For example, *Våagnes and Amundsen* [1993] documented 1.7 km of denudation resulting from surface uplift of Spitzbergen since 10 Ma, and *van der Beek* [1995] argued for ~1.5 km of tectonic uplift in southern Norway since 30 Ma.

[49] Two major tectonic events affected the European plate margins during the Cenozoic: the Alpine orogeny, and ridge-push forces in the NE Atlantic. *Doré et al.* [1999] showed that for the Cenozoic there was a reasonable correspondence between phases of Alpine shortening on one margin of the plate with a slowing of spreading and hence increased shortening along the opposite, passive margin.

[50] *Ziegler* [1988] suggests that inversion in NW Europe was due to Alpine compression, culminating in a late Alpine docking event (Tethyan closure) of Miocene age. *Blundell* [2002] correlated Cenozoic inversion events in southern

Britain with far-field Alpine tectonics. He showed that Alpine stresses were only transmitted into the foreland when they could not be accommodated within the orogen or flexural foreland basin. *Ford and Lickorish* [2004] provided evidence for continued flexural subsidence from Oligocene to late Miocene in the North Alpine Foreland basin. Subsequent alpine deformation and uplift is seen in the Paris Basin, the Rhine Graben, the Lulworth folds of southern Britain and Late Tertiary inversion in the North Sea. This event may also have played a major role in inversion and exhumation of the Irish Sea region.

[51] A number of authors have correlated Oligocene-Miocene inversions of Norwegian basins with plate reorganization in the adjacent ocean [e.g., *Brekke and Riis*, 1987; *Doré and Lundin*, 1996]. *Cloetingh et al.* [1990] argued for deformation and uplift in NW Europe because of Pliocene plate reorganization and changes in intraplate stress; however, much of the denudation and associated sedimentation along the NW European margin began in pre-Pliocene time. In general, observed occurrences of Cenozoic crustal shortening in NW Europe is on the scale of 1–2%, e.g., western Ireland [*Dewey*, 2000], basin inversions offshore Norway [*Doré et al.*, 1999], and compression along the Faroe-Rockall Plateau [*Boldreel and Andersen*, 1998]. These magnitudes of shortening cannot account for the total amount of Cenozoic exhumation inferred from vitrinite reflectance profiles and apatite fission track analysis [*Brodie and White*, 1995]. Therefore a number of authors have related Cenozoic denudation to regional uplift [e.g., *Lovell and White*, 1997; *Jones et al.*, 2002].

[52] *Våagnes and Amundsen* [1993] called on secondary mantle convection and underplating as an uplift mechanism. *Rohrman and van der Beek* [1996] invoked asthenospheric diapirism resulting from a Rayleigh-Taylor instability, arguing that the initiation of Cenozoic tectonic and magmatic activity coincided with the imposition of hot Iceland plume asthenosphere beneath the margin of the European continental lithosphere at ~30 Ma. *Rohrman and van der Beek* [1996] indicated that northern Britain and Ireland should be underlain by an asthenospheric diapir, potentially causing both transient and permanent tectonic rock uplift.

[53] The importance of the ridge-push mechanism is reinforced by post-Oligocene development of broad wavelength, low amplitude folds in the Traill Ø region of East Greenland [*Price et al.*, 1997], and by early Eocene to Miocene folding of Paleocene lavas on the Faroe-Rockall Plateau, adjacent to the Aegir Ridge [*Boldreel and Andersen*, 1998] (Figure 9). However, compression on the flank of the Icelandic mantle plume also needs to be considered as a possible mechanism for the development of these structures.

[54] *Dewey* [2000] noted that the predominantly NW-SE trends of Paleocene dikes in Scotland, northern Ireland, and the Irish Sea swing to E-W and NE-SW trends in western Ireland, suggesting that western Ireland is more distant from the flanks of the Iceland plume head. *Dewey* [2000] also noted that NW Europe was caught in the jaws of a compressional vice from 22 to 9 Ma, created both by rapid convergence between Africa and Europe and by North

Atlantic ridge-push forces. He concluded that Cenozoic uplift and deformation in western Ireland was driven by rift-shoulder uplift in the northeastern Atlantic and subsequent ridge-push effects.

[55] In summary, Cenozoic tectonic deformation and rock uplift on the NW European margin is most probably due to intraplate stresses driven from interaction between ridge-push and far-field Alpine compression [e.g., Roberts, 1989; Doré and Lundin, 1996]. This link is strengthened by modeling of the European stress field showing that the present-day field, mainly NW-SE maximum horizontal compression [Zoback, 1992; Müller et al., 1992, 1997], is caused by a combination of ridge-push and collisional foreland stresses [Gölke and Coblenz, 1996]. Enhanced denudation due to Pliocene-Pleistocene glaciations and isostatic rebound during interglacial periods was also an important factor in shaping the margin, but was probably of more local, rather than regional, significance [Doré et al., 1999; Solheim et al., 1996].

[56] In SE Ireland, exhumation began post-early Eocene [Cunningham et al., 2003] as part of a regional cooling event that affected the Irish landmass at this time [Allen et al., 2002]. Cunningham et al. [2003] showed that the major rivers of SE Ireland are out of equilibrium with the present topography in a manner that is consistent with the locations,

magnitudes, and vertical displacements on faults, that were probably reactivated syn- or post-early Eocene. Episodes of shortening during the Oligocene in NE Ireland and the Irish Sea, and during post-Oligocene time in SE Ireland, show that local inversions were partially responsible for rock uplift and denudation during the Cenozoic. This pattern is consistent within the regional tectonic framework of the NW European margin (Figure 9). However, NE Atlantic ridge-push forces appear to be most dominant geodynamic mechanisms driving post-Oligocene inversion in rock uplift in SE Ireland.

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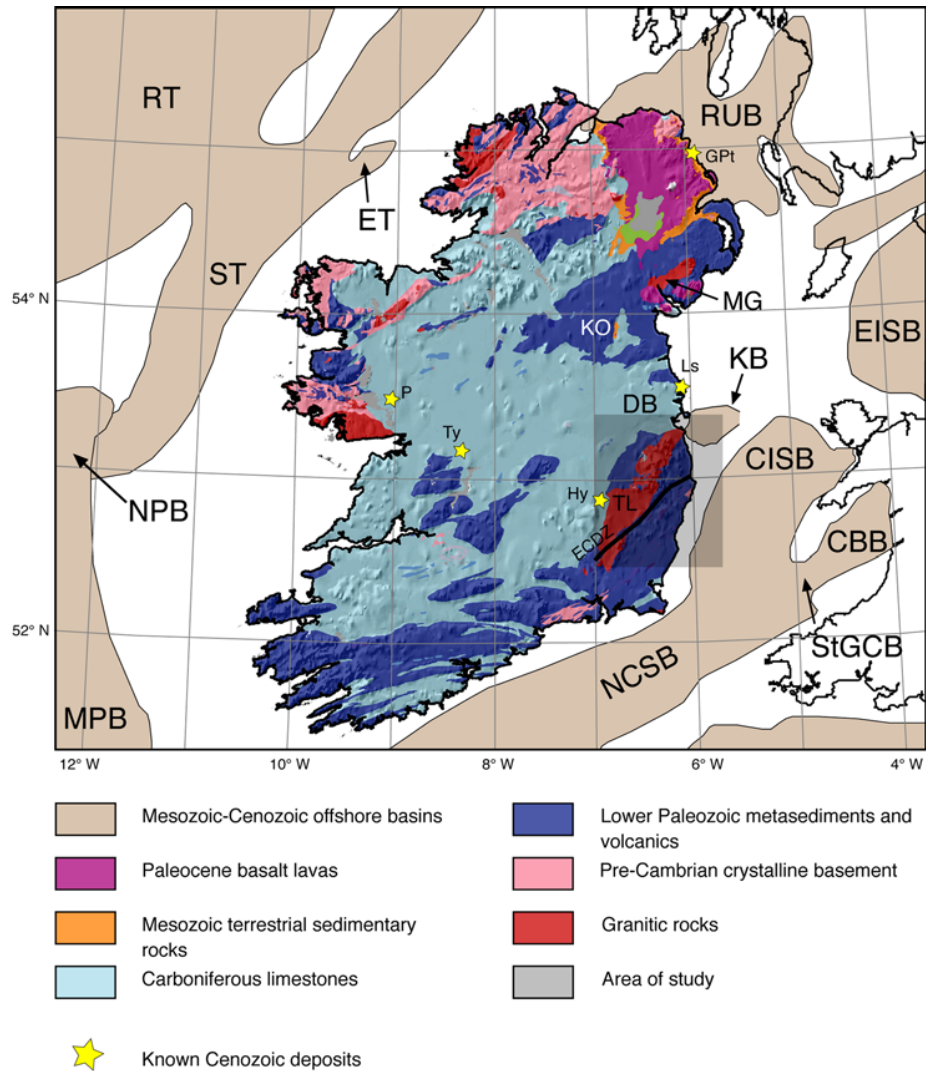


Figure 1. (a) Onshore geology and topography of Ireland and major Mesozoic-Cenozoic offshore basins. Study area is shown by the gray box in SE Ireland. Abbreviations area as follows: CBB, Cardigan Bay Basin; CISB, Central Irish Sea Basin; DB, Dublin Basin; ECDZ, East Carlow Deformation Zone; EISB, East Irish Sea Basin; ET, Erris Trough; Gpt, Garron Point; Hy, Hollymount; KO, Kingscourt Outlier; Ls, Loughshinney; MG, Mourne Granite; MPB, Main Porcupine Basin; NCSB, North Celtic Sea Basin; NPB, North Porcupine Basin; P, Pollnahallia; RUB, Rathlin/Ulster Basin; RT, Rockall Trough; StGCB, St. George's Channel Basin; ST, Slyne Trough; TL, Tullow Lowland; Ty, Tynagh.

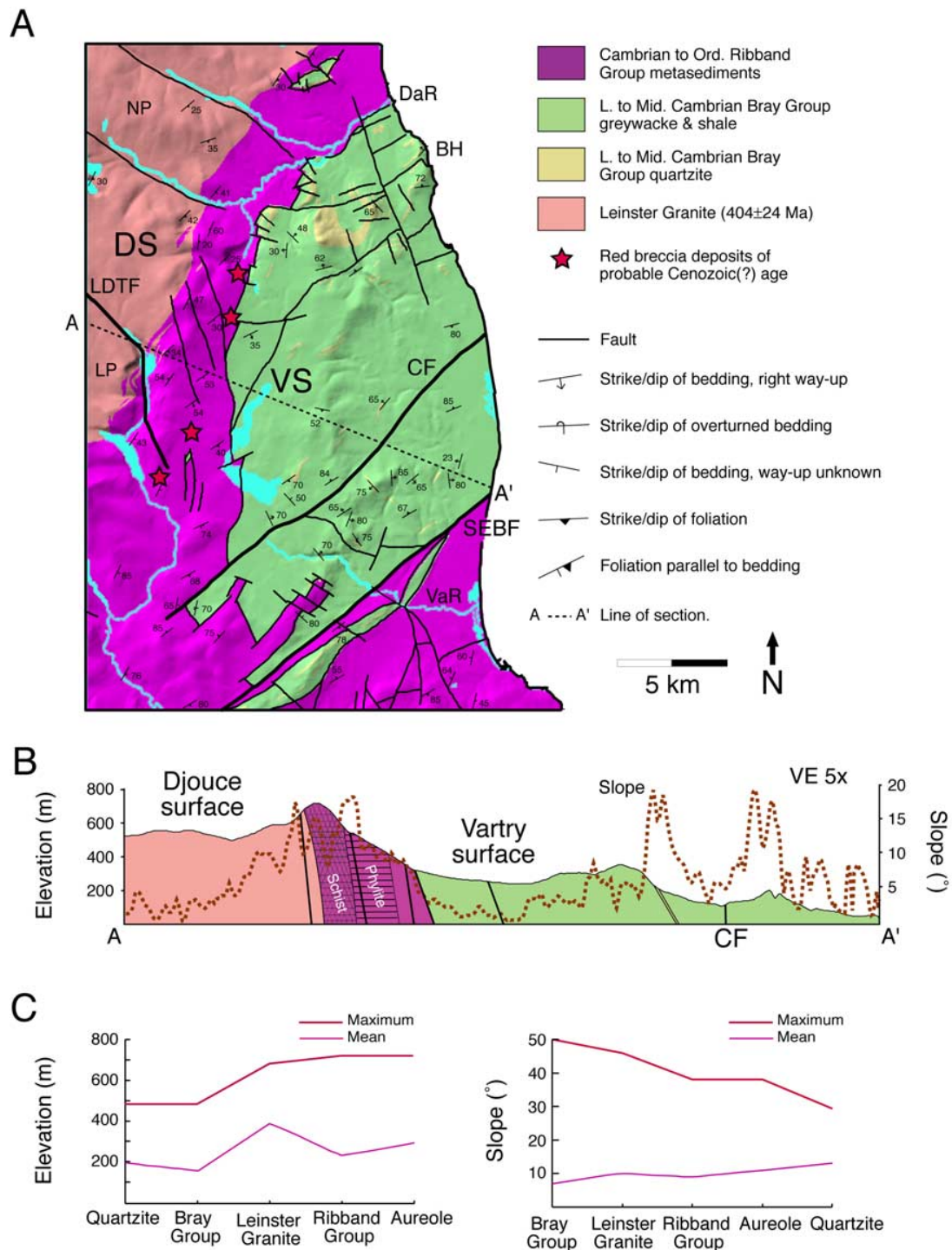


Figure 3. (a) Bedrock geology shaded relief image of SE Ireland. Abbreviations area as follows: BH, Bray Head; CF, Callowhill Fault; DaR, River Dargle; DS, Djouce surface; LDTF, Lough Dan-Tay Fault; SEBF, Southeastern Boundary Fault; VaR, Vartry River; VS, Vartry surface. (b) WNW-ESE trending cross section showing elevation, topographic slope (dashed line, calculated from the digital elevation model using 3×3 window of 50 m cells), and bedrock geology. (c) Mean and maximum values of elevation and slope for the various bedrock lithologies. Aureole rocks and quartzite are included within Ribband and Bray Groups, respectively, but are also shown separately as a guide to their effects on slope.

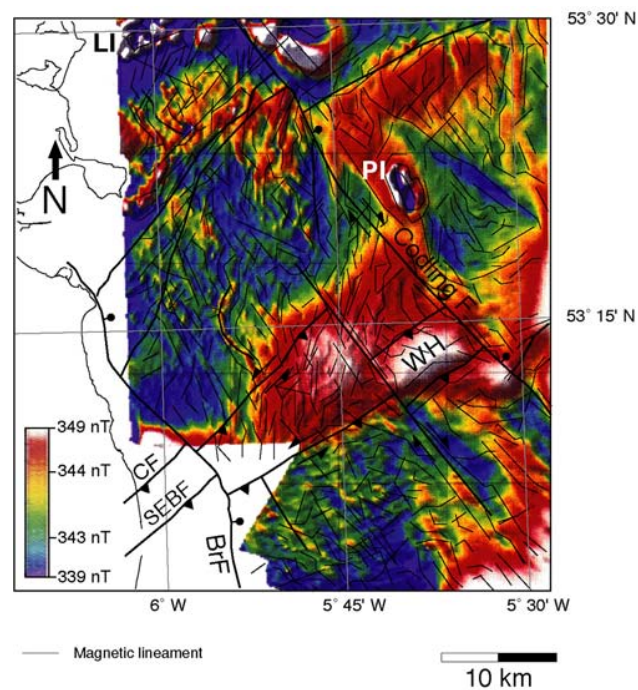


Figure 7. Total magnetic intensity image of the Kish Bank Basin. Image area is the same as in Figures 6a and 6b, and fault traces are overlain for comparison. Magnetic intensity is reduced to pole with a pseudo depth slice filter applied. The filter accentuates high-frequency, low-amplitude shallow magnetic structure. Note trends in major magnetic lineaments indicated by thin lines (see section 4.2 for full discussion). Abbreviations are as follows: BrF, Bray Fault; CF, Callowhill Fault; LI, Lambay Island; PI, Intrusion; SEBF, Southeastern Boundary Fault; WH, Wicklow High.